Interdecadal Variations of the East Asian Winter Monsoon and Their Association with Quasi-Stationary Planetary Wave Activity

LIN WANG, RONGHUI HUANG, LEI GU, WEN CHEN, AND LIHUA KANG

Center for Monsoon System Research, Institute of Atmospheric Physics, Chinese Academy of Sciences, Beijing, China

(Manuscript received 11 December 2008, in final form 9 March 2009)

ABSTRACT

Interdecadal variations of the East Asian winter monsoon (EAWM) and their association with the quasi-stationary planetary wave activity are analyzed by using the 40-yr European Centre for Medium-Range Weather Forecasts Re-Analysis dataset and the National Centers for Environmental Prediction–National Center for Atmospheric Research reanalysis dataset. It is found that the EAWM experienced a significant weakening around the late 1980s; that is, the EAWM was strong during 1976–87 and became weak after 1988. This leads to an obvious increase in the wintertime surface air temperature as well as a decrease in the frequency of occurrence of cold waves over East Asia. The dynamical process through which the EAWM is weakened is investigated from the perspective of quasi-stationary planetary waves. It is found that both the propagation and amplitude of quasi-stationary planetary waves have experienced obvious interdecadal variations, which are well related to those of the EAWM. Compared to the period 1976–87, the horizontal propagation of quasi-stationary planetary waves after 1988 is enhanced along the low-latitude waveguide in the troposphere, and the upward propagation of waves into the stratosphere is reduced along the polar waveguide. This results in a weakened subtropical jet around 40°N due to the convergence of the Eliassen–Palm flux. The East Asian jet stream is then weakened, leading to the weakening of the EAWM since 1988. In addition, the amplitude of quasi-stationary planetary waves is significantly weakened around 45°N, which is related to the reduced upward propagation of waves from the lower boundary after 1988. This reduced amplitude may weaken both the Siberian high and the Aleutian low, reduce the pressure gradient in between, and then weaken the EAWM. Further analyses indicate that zonal wavenumber 2 plays the dominant role in this process.

1. Introduction

As one of the most active components of the global climate system, the East Asian winter monsoon (EAWM) is an important climate feature over East Asia in boreal winter. The most prominent surface feature of the EAWM is the strong northwesterly along the east flank of the Siberian high. This northwesterly flow bifurcates south of Japan, with one branch turning eastward toward the subtropical western and central Pacific and the other flowing along the coast of East Asia into the South China Sea and the Indo-China Peninsula (e.g., Lau and Li 1984; Ding 1994; Chen et al. 2000, 2005). The variability of the EAWM may exert large social and economic impacts in many Asian countries. For example, strong EAWM can frequently bring severe cold waves/snowstorms in Korea, Japan, and northern China; persistent cooling disasters in South China; and severe flooding in Southeast Asian countries (e.g., Ding 1994; Huang et al. 2003; Chang et al. 2006). In January 2008, prolonged cooling and severe snowstorms attacked most of China. One hundred, twenty-nine people died and the economic losses reached up to ~150 billion yuan (~21 billion U.S. dollars; NCC/CMA 2008). Thus, it is very important and necessary to understand variability of the EAWM.

The pioneering studies of the EAWM can be traced back to the 1950s when some Chinese meteorologists began to investigate the circulation characteristics and precursors of the East Asian cold wave through case analyses (e.g., Tao 1957; Tao et al. 1965). Since the 1980s, the research interests were mainly on the relationship between cold waves and the Siberian high through statistical and diagnostic analyses (e.g., Ding and Krishnamurti 1987; Ding 1990; Ding et al. 1991). In recent years, more
and more attention has been paid to the interannual variability of the EAWM and related driving factors (e.g., Chen et al. 2000; Gong et al. 2001; Jhun and Lee 2004; Wang et al. 2009b), and many valuable results have been obtained. For example, Chen et al. (2000) defined an EAWM intensity index by averaging the low-level meridional wind in a certain area over coastal China, through which the impact of El Niño–Southern Oscillation on the EAWM was clearly revealed, with weak (strong) EAWM during El Niño (La Niña) winters. Wang et al. (2009b) investigated the interannual variations of the EAWM pathway and the impact of North Pacific sea surface temperature (SST) on such variations by analyzing the 500-hPa East Asian trough.

On the other hand, the impact of the Arctic Oscillation/North Atlantic Oscillation (AO/NAO) on the EAWM has been widely discussed since the concept of the AO (Thompson and Wallace 1998) was proposed. Wu and Huang (1999) first noted that the EAWM tends to be weak during the positive phase of the AO. Gong et al. (2001) pointed out that the AO exerts its influence on the EAWM through the Siberian high, whereas Wu and Wang (2002) argued that the influence of the AO on the EAWM was independent of the Siberian high. Since variability of the AO can significantly modulate the propagation of quasi-stationary planetary waves between the troposphere and the stratosphere (Chen et al. 2000, 2003; Chen and Huang 2005), Chen et al. (2005) examined the interannual AO–EAWM relationship from the perspective of the quasi-stationary planetary wave activity (PWA). They found that during the positive phase of the AO, more quasi-stationary planetary waves propagate from high latitudes to lower latitudes in the troposphere instead of propagating vertically to the stratosphere. This is accompanied by a smaller perturbation in the polar vortex, which tends to be colder and stronger. Meanwhile, the Siberian high and the Aleutian low are both weakened significantly. This reduces the pressure gradient and the northeasterly along the coasts of East Asia, leading to a weak EAWM. This explanation emphasizes not only the role of the Siberian high but also that of the Aleutian low in the variability of the EAWM.

In addition to the interannual variability, the EAWM is also characterized by obvious variability on interdecadal time scales (Shi 1996; Jhun and Lee 2004). Observational studies reveal that the EAWM was significantly weakened after the late 1980s (e.g., Huang and Wang 2006; Kang et al. 2006), which is accompanied by frequently warm winters in China (e.g., Wang and Ding 2006; Kang et al. 2006) and weakened sea surface wind stress along the coasts of East Asia (e.g., Cai et al. 2006). This weakening of the EAWM since the 1980s is believed to be associated with variations of the AO, which may be accounted for by the excessive autumn snowfall over the northeastern Eurasian continent (Jhun and Lee 2004). Although supported by model results (Jhun and Lee 2004), this explanation does not make clear how the almost zonally symmetric AO induces a weakened EAWM since 1988. Therefore, the circulations and internal atmospheric processes associated with the interdecadal variations of the EAWM will be examined in this study. The effort will be made to understand how the EAWM is weakened on the interdecadal time scale from the perspective of the quasi-stationary planetary wave activity.

The datasets and analysis methods used in this study are described in section 2. The interdecadal variations of the EAWM and their influence on the East Asian climate are then illustrated in section 3. Section 4 presents the climatology of quasi-stationary planetary waves, which serves as a basis for section 5. Section 5 then documents the interdecadal variations of the quasi-stationary planetary waves and their relationship with those of the EAWM. Finally, a summary and discussion is given in section 6.

2. Data and methods

Atmospheric data used in this study include the monthly-mean 40-yr European Centre for Medium-Range Weather Forecasts (ECMWF) Re-Analysis (ERA-40) data (Uppala et al. 2005) covering the period from September 1957 to August 2002. This dataset has a horizontal resolution of approximately 1.125° × 1.125° and extends from 1000 to 1 hPa with 23 vertical pressure levels. Atmospheric data also include the reanalysis data from the National Centers for Environmental Prediction–National Center for Atmospheric Research (NCEP–NCAR), which covers the period from January 1948 to present (Kalnay et al. 1996). This dataset has a horizontal resolution of 2.5° × 2.5° and extends from 1000 to 10 hPa with 17 vertical pressure levels. The NCEP–NCAR reanalysis data are used to validate and compare with the results derived from ERA-40. For the NCEP–NCAR reanalysis, we only adopt the time period that overlaps with ERA-40 data. In addition, the China cold wave dataset collected by the National Climate Center, China Meteorological Administration (NCC/CMA) is employed to represent the frequency of cold waves associated with the EAWM in China.

As a diagnostic tool, the Eliassen–Palm (EP) flux, which is a measure of the wave activity propagation (Andrews et al. 1987), is calculated in spherical geometry. Its divergence owing to planetary waves indicates the eddy forcing of the zonal-mean flow. The components of EP flux \( F \) and its divergence \( D_F \) are
\[ F_y = -\rho a \cos \phi \overline{u' v'}; \quad F_z = \rho a \cos \phi \frac{Rf}{HN^2} \overline{u'T'}; \]
\[ D_F = \frac{\mathbf{V} \cdot \mathbf{F}}{\rho a \cos \phi}, \]

where \( \rho \) is air density, \( a \) the radius of the earth, \( \phi \) the latitude, \( f \) the Coriolis parameter, \( R \) the gas constant, \( N \) the buoyancy frequency, \( H \) a constant-scale height (about 7 km), \( u \) and \( v \) are the zonal and meridional wind, and \( T \) is the air temperature; primes denote zonal deviation, and overbars denote zonal average. Here \( N \) is calculated from the temperature data. By expanding the monthly-mean fields into their zonal Fourier harmonics, the zonal wavenumbers 1 through 3 are used to represent the quasi-stationary planetary wave. Then they are used to calculate the EP flux and its divergence, similar to the method of Chen et al. (2003). In addition, the zonal Fourier harmonics of geopotential height with wavenumbers 1–3 are used in calculating the amplitudes of quasi-stationary planetary waves following the method of van Loon et al. (1973).

The time period analyzed is from 1957 to 2002. Seasonal means are considered throughout this paper and are constructed from the monthly means by averaging values in December–February (DJF), which results in 45 winters (1957–2001). One exception is the number of cold waves, which is the average for the extended winter [November–April (NDJFMA)] owing to the availability of data. Here the winter of 1957 refers to the boreal 1957/58 winter. In this study, the climatology mean is defined as the average from 1971 to 2000, and the anomaly is the departure from this climatology mean.

3. Interdecadal variability of the EAWM and its impact on the East Asian winter climate

The EAWM is characterized by obvious interannual variability as well as interdecadal variability (e.g., Shi 1996; Jhun and Lee 2004; Huang and Wang 2006). The interdecadal variability and associated climate features of the EAWM are relatively less documented. Thus, in this section, we first describe briefly the interdecadal variability of the EAWM and its impact on the winter-time climate over East Asia.

a. The EAWM index and the interdecadal variability of the EAWM

In boreal winter, the most significant sea level pressure (SLP) systems over East Asia are the Siberian high and Aleutian low (Fig. 1). The strong pressure gradient between these two systems favors a northerly wind along the east coast of the Eurasian continent, which is one of the most important surface features of the EAWM (Zhang et al. 1997; Chen et al. 2000). Both interannual and interdecadal variations of the EAWM are closely related to the anomalies of these two systems and the associated pressure gradient. Therefore, based on the literature of Wu and Wang (2002), an EAWM index that mainly concerns the Siberian–Aleutian pressure contrast is defined as follows:

\[ I_{EAWM} = \text{norm} \left[ \sum_{\text{lat}=40^\circ \text{N}}^{70^\circ \text{N}} (\text{norm}(p_{110^\circ \text{E}}) - \text{norm}(p_{160^\circ \text{E}}))_{\text{lat}} \right], \]

where \( p_{110^\circ \text{E}} \) and \( p_{160^\circ \text{E}} \) represent the winter mean SLP at each grid point along 110° and 160°E, respectively, and
norm denotes the normalization. Compared with their definition, in which the defined latitudes cover 20°–70°N, the area of the present study is limited to 40°–70°N, where the main centers of both the Siberian high and the Aleutian low are located (Fig. 1). Following this definition, the EAWM is anomalously strong when $I_{\text{EAWM}}$ is positive, and vice versa.

Figure 2 illustrates the normalized winter mean EAWM index from 1957 to 2001. Comparison of Fig. 2a with Fig. 2b reveals that the index calculated from the ERA-40 data (Fig. 2a) is similar to that from the NCEP–NCAR reanalysis data (Fig. 2b). Both indices are characterized by obvious interannual and interdecadal fluctuations, with a correlation coefficient of 0.802 that exceeds the 99% confidence level. Although there are some discrepancies in the mean value before and after 1988 between Figs. 2a and 2b, both indices show a clear transition of the EAWM intensity around 1988 with strong EAWM before 1988 and weak EAWM afterward. Based on a two-sided Student’s $t$ test, the mean difference between the two periods exceeds the 95% confidence level in both the ERA-40 (Fig. 2a) and the NCEP–NCAR datasets (Fig. 2b). This result is consistent with that reported in previous studies (e.g., Jhun and Lee 2004; Huang and Wang 2006). In the following, we choose 1976–87 and 1988–2001 to represent the strong and weak EAWM periods, respectively, and analyze the climate features and associated circulation and planetary wave activity during these two periods. The reason why we only consider the period after 1976 is that a significant climate shift occurred around 1976 (Trenberth and Hurrell 1994) and the East Asian climate system also changed around that time (e.g., Wang et al. 2007, 2008, 2009a). So as to avoid the possible influence of this shift on our results, we only select the years after 1976 for investigation.

b. Impact of the EAWM interdecadal weakening on the East Asian winter climate

The weakening of the EAWM since 1988 has an important impact on the winter climate in East Asia. Figure 3 presents the winter mean SLP and surface air temperature (SAT) anomalies for the two periods, 1976–87 and 1988–2001, and their difference. It reveals that during the strong EAWM period (1976–87), the Aleutian low is intensified, so is the western part of the Siberian high (Fig. 3a). This SLP distribution favors stronger northerly wind around the eastern part of Eurasian continent and causes anomalous cooling over large areas of East Asia (Fig. 3b). During the weak EAWM period (1988–2001), the SLP anomalies are generally

\[ \text{FIG. 3. The (a) SLP and (b) surface air temperature anomalies averaged for the 1976–87 winter based on the ERA-40 dataset. (c),(d) As in (a),(b), but for the 1988–2001 winter. Contour intervals are 0.5 hPa in (a),(c) and 0.5°C in (b),(d).} \]
opposite to those of strong EAWM period (Fig. 3c), leading to a clear warming over East Asia (Fig. 3d). The cold wave, which usually originates from the Siberian area, is one of the most important synoptic features of the EAWM (e.g., Tao 1957; Zhang et al. 1997). It can cause an abrupt and large drop of temperature in China, Korea, and Japan and strong convection over the Maritime Continent (e.g., Chang et al. 1979; Lau and Chang 1987). Based on the criteria of NCC/CMA, the cold waves can further be classified into general and strong ones based on their intensity. Figure 4 presents the number of cold waves in China for the extended winters during 1957–2001, recorded in the NCC/CMA cold wave almanac.

The cold wave, which usually originates from the Siberian area, is one of the most important synoptic features of the EAWM (e.g., Tao 1957; Zhang et al. 1997). It can cause an abrupt and large drop of temperature in China, Korea, and Japan and strong convection over the Maritime Continent (e.g., Chang et al. 1979; Lau and Chang 1987). Based on the criteria of NCC/CMA, the cold waves can further be classified into general and strong ones based on their intensity. Figure 4 presents the number of cold waves in China for the extended winters during 1957–2001, recorded in the NCC/CMA cold wave almanac.

4. The climatology of the quasi-stationary planetary waves

Before we investigate the interdecadal variations of the quasi-stationary planetary waves associated with those of the EAWM, it is necessary to present the climatology of the quasi-stationary planetary waves to serve as a background for the following analyses. van Loon et al. (1973) first calculated the latitude–altitude distribution of amplitudes and phases of quasi-stationary planetary waves for wavenumbers 1 through 3 over the Northern Hemisphere in January. However, no updated work has been reported since then. On the other hand, the time period that van Loon et al. (1973) considered is 1964–70. The mean state calculated from this short period is not suitable for investigating the problems concerning the interdecadal variability. Therefore, in this section, we first present the climatology—including the amplitude and propagation—of quasi-stationary planetary waves based on the ERA-40 and NCEP–NCAR reanalysis datasets.

Figures 5a–c illustrate the long-term mean wintertime amplitudes of quasi-stationary planetary wavenumber 1–3 based on the ERA-40 data. The maximum amplitude of wavenumber 1 is located around 10 hPa, 65°N in the stratosphere with the value exceeding 750 gpm (Fig. 5a). In addition, it has a secondary peak at the low-latitude tropopause around 150 hPa, 20°N (Fig. 5a). The maximum amplitude of wavenumber 2 exceeds 250 gpm and is located around 10 hPa, 60°N in the stratosphere (Fig. 5b). For wavenumber 3, the amplitude has a maximum (over 100 gpm) at ~250 hPa, 50°N in the troposphere and a second peak at ~200 hPa, 20°N (Fig. 5c). These results compare well with those calculated for January of 1964–70 by van Loon et al. (1973), except for slightly larger amplitude in the stratosphere for wavenumber 1.

Figures 5d–f show the climatology of wintertime amplitudes of wavenumber 1–3 based on the NCEP–NCAR reanalysis data. Compared with those from ERA-40 data (Figs. 5a–c), the amplitudes calculated from the NCEP–NCAR reanalysis data have almost the same spatial distributions and very close values (Figs. 5d–f). The only important difference is in the maximum amplitudes in the stratosphere, where the values based on the NCEP–NCAR reanalysis are smaller than those based on ERA-40. This is most clear for wavenumber 2, whose amplitude is over 250 gpm in ERA-40 (Fig. 5b) but only exceeds 200 gpm in the NCEP–NCAR reanalysis (Fig. 5e). This discrepancy may be related to the higher model top and higher vertical resolution of the ERA-40 assimilating system (Uppala et al. 2005) since the vertical resolution is very important for the vertical structure of planetary waves (e.g., Giorgetta et al. 2002). Nevertheless, the distributions of wave amplitudes (Figs. 5) correspond well with the propagation of quasi-stationary planetary waves (e.g., Chen et al. 2003, their Figs. 1b–d), since the disturbances are caused by the incoming wave propagations.
In addition to the amplitudes, it is also helpful to look at the climatology of wave activity. Figure 6 presents the climatology of the winter-mean EP flux for the sum of wavenumbers 1–3 based on ERA-40 and the NCEP–NCAR reanalysis. It is clear that planetary waves propagate upward from the lower boundary over midlatitudes and split into two branches in the upper troposphere with one branch propagating into the stratosphere along the polar waveguide (Dickinson 1968; Matsuno 1970) and the other propagating equatorward along the low-latitude waveguide (Huang and Gambo 1983) in the troposphere. This diagram compares well with the theoretical and simulated results of planetary wave propagation (Matsuno 1970; Huang and Gambo 1982) and also to the previous observations (Dunkerton and Baldwin 1991; Chen and Huang 1999, 2002). The point is that the upward propagation of planetary waves in the ERA-40 dataset (Fig. 6a) is stronger than that in the NCEP–NCAR dataset (Fig. 6b). This is consistent with the previous amplitude analyses and implies that it is more suitable in some sense to use the ERA-40 dataset to analyze the variations of the planetary waves associated with the EAWM. Therefore, only the results derived from the ERA-40 dataset are investigated in the following section.

5. The interdecadal anomalies of the quasi-stationary planetary waves and their associations with the EAWM

To quantify the quasi-stationary planetary wave activity, a PWA index is defined after the method of Chen et al. (2003) as

\[
I_{\text{PWA}} = \text{norm}[(\nabla \cdot \mathbf{F})_A - (\nabla \cdot \mathbf{F})_B],
\]

where \((\nabla \cdot \mathbf{F})_A\) and \((\nabla \cdot \mathbf{F})_B\) are the divergence of EP fluxes for the sum of wavenumbers 1–3 at point A (49.906°N, 500 hPa) and point B (39.813°N, 300 hPa) and norm denotes the normalization. Following this definition, more planetary waves are found to propagate horizontally to the subtropical tropopause along the low-latitude waveguide in high PWA years (Chen et al. 2003; see also Fig. 8b) and vice versa.
Figure 7 shows the PWA index for the 1957–2001 winters. The correlation coefficient between the PWA index and EAWM index is 0.41 for the 45 winters, exceeding the 99% confidence level. In addition, the PWA index is in its low phase during 1976–87 and in its high phase during 1988–2001 (Fig. 7), with the difference exceeding the 99% confidence level. This transition is consistent with that of the EAWM index around 1988 (Fig. 2). These results imply a close relationship between planetary wave activity and the intensity of EAWM on an interdecadal time scale. To reveal the interdecadal relationship more clearly, we compare the characteristics of planetary waves for the low (1976–87) and the high (1988–2001) PWA index periods in the following.

Figure 8 illustrates the anomalous EP flux and its divergence (sum of wavenumber 1–3) of the quasi-stationary planetary waves for the low PWA period (1976–87), the high PWA period (1988–2001), and their difference. During the low PWA period, there is anomalous horizontally poleward propagation of waves along the low-latitude waveguide in the upper troposphere and anomalous upward propagation of waves along the polar waveguide north of 65°N (Fig. 8a). This is accompanied with anomalous EP flux divergence around 40°N, 250 hPa and convergence in the middle troposphere north of that (Fig. 8a). Consequently, the westerly wind is intensified (weakened) around 35°N (55°N) (Fig. 9a) as convergence of EP flux can lead to the deceleration of zonal winds, and vice versa (Andrews et al. 1987). During the high PWA period, the horizontally equatorward wave propagation along the low-latitude waveguide is significantly intensified, and the upward propagation of waves along the polar waveguide does not change much (Fig. 8b). Therefore, the associated anomalous EP flux divergence (Fig. 8b) and zonal winds (Fig. 9b) are generally reversed compared to those during the low PWA period. The differences between the two periods are quite clear, with strong equatorward propagating EP flux along the low-latitude waveguide and significant EP flux convergence around 40°N, 300 hPa (Fig. 8c). The corresponding zonal-mean zonal wind difference exhibits a clear dipole with the node located at ~45°N (Fig. 9c). This picture resembles the traditional AO pattern (e.g., Thompson and Wallace 2000) but with the
significant signals mostly trapped in the troposphere. It suggests that the most significant change of zonal-mean zonal winds is in the troposphere on the interdecadal time scale—in contrast to that on the interannual time scale (e.g., Thompson and Wallace 2000). Therefore, accompanying more equatorward propagation of quasi-stationary planetary waves along low-latitude wave-guide, the subtropical jet is significantly weakened around 35°N, 300 hPa during the high PWA period (Fig. 9c).

Although the weakened zonal-mean subtropical jet after 1988 may imply a weakened EAWM (e.g., Yang et al. 2002; Jhun and Lee 2004), it is necessary to examine how the atmospheric circulation over East Asia varies with the changes in planetary wave activity since the flow in the Northern Hemisphere is not zonally symmetric. Figure 10 then shows the 300-hPa zonal wind patterns over East Asia for the periods 1976–87 and 1988–2001 and their difference. The upper-atmospheric circulation over extratropical East Asia is dominated by

![Fig. 8. Anomalies of the EP-flux cross section (vectors) and its divergence (contour) for the sum of zonal wave-numbers 1 to 3 based on the ERA-40 dataset for the winters of (a) 1976–87 and (b) 1988–2001 and (c) their difference (1988–2001 minus 1976–87). Contour intervals are 3 × 10^{-5} m^2 s^{-2}. EP fluxes (m^2 s^{-2}) are scaled by the inverse of the air density. Light (dark) shading in (c) indicates the 95% (99%) confidence level.](image_url)

![Fig. 9. Anomalies of zonal-mean zonal wind based on the ERA-40 dataset for the winters of (a) 1976–87 and (b) 1988–2001 and (c) their difference (1988–2001 minus 1976–87). Contour intervals are 0.5 m s^{-1}. Light (dark) shading in (c) indicates the 95% (99%) confidence level.](image_url)
the East Asian westerly jet stream with the maximum over the south of Japan (Figs. 10a and 10b). In the period of high wave activity (1988–2001), the jet stream becomes weaker, especially in the center and the downstream portion (Fig. 10c). This suggests that the more equatorward propagation of planetary waves can lead to a weakened East Asian jet stream and, hence, a weakened EAWM during the high wave activity period (1988–2001), as the weakened East Asian jet stream and associated intensified polar jet is a typical feature of weak EAWM (e.g., Jhun and Lee 2004).

In addition to changes in the wave-convergence-induced jet stream, changes in the amplitude of planetary waves also contribute to the weakening of the EAWM. Figure 11 shows the anomalous amplitude of wavenumber 1–3 for the low (1976–87) and high (1988–2001) wave activity periods, and their difference. It reveals that, during the low wave activity period, the amplitude of planetary waves is anomalously high around 45° and 75°N and low around 30° and 60°N (Fig. 11a). This signal has its peaks in the middle and upper troposphere with the anomaly centers around 45° and 60°N extending to the surface. Among these four anomaly centers, the increased amplitude around 45°N has the closest relationship with the EAWM since it is close to the latitude around which the Siberian high and Aleutian low are located (Fig. 1). The changed amplitude of planetary waves is accompanied by strengthening of both the Siberian high and the Aleutian low (e.g., Figs. 3a,c; see also Fig. 13), leading to changes in the intensity of the EAWM (e.g., Fig. 3b,d). During the high wave activity period, the amplitude anomalies are generally reversed (Fig. 11b). Therefore, the wave amplitude is significantly weakened around 45°N after 1988 (Fig. 11c). However, it is interesting to note that, although the maximum anomaly around 45°N is in the upper troposphere, the significant signal is mainly confined to the lower atmosphere near the surface (Fig. 11c), implying that the changes at or near the surface are most important. Watanabe and Nitta (1999) suggested that the autumn Eurasian snow cover may play important roles for the decadal change of East Asian climate around 1988, so we examined the difference of autumn [September–November (SON)] snow depth between 1988–2001 and 1976–87. It reveals that the autumn snow depth over large areas of northeastern Eurasia, as well as North America, decreased significantly after 1988 (Fig. 12). This may decrease the land–sea thermal contrast over mid and high latitudes and reduce the amplitude of planetary waves around 45°N.

One issue that should be addressed is how the planetary wave amplitude changes are linked to the changes of the EAWM intensity. We investigated the regimes of individual zonal wavenumbers from 1 to 3 and found that the zonal-mean distribution of amplitude anomalies (Fig. 11) are mainly contributed from zonal wavenumber 2, especially for the center around 45°N (not shown). The weakened amplitude in this region may in turn weaken the Siberian high and Aleutian low, both of which are located around the same latitude. This feature can be seen more clearly in the wavenumber 2 pattern of the SLP field (Fig. 13). It reveals that the maximum

---

**Fig. 10.** The 300-hPa zonal wind based on the ERA-40 dataset for the winters of (a) 1976–87 and (b) 1988–2001 and (c) their difference (1988–2001 minus 1976–87). Contour intervals are 5 m s⁻¹ in (a), (b) and 2 m s⁻¹ in (c). Light (dark) shading in (c) indicates the 95% (99%) confidence level.
The amplitude of wavenumber 2 appears near 50°N, with high pressure controlling most of the continental area and low pressure dominating over the oceans (Figs. 13a,b). This wave pattern coincides well with the features of climatological pressure systems including the Siberian high, the Aleutian low (e.g., Fig. 1), the Canadian high, and the Icelandic low. During the high wave activity period (1988–2001), the amplitude of the high and low pressures shown in Figs. 13a and 13b decreases significantly. It can be seen by comparing Fig. 13c with Figs. 3a and 3c that the contribution from wavenumber 2 accounts for more than 75% and 50% of the changes in the Siberian high and the Aleutian low, respectively. On the contrary, there is no significant contribution by wavenumbers 1 and 3 to the changes of the Siberian high and Aleutian low (not shown). Therefore, the variations of wavenumber 2 play an important role in the weakening of the EAWM since 1988. This result also coincides well with the change of Eurasian and North American snow cover (Fig. 12) as the wavenumber 2 pattern in the Northern Hemisphere is closely associated with the thermal contrast between land and oceans (Huang and Gambo 1982).

6. Summary and discussion

Based on the ERA-40 and the NCEP–NCAR reanalysis datasets, we investigated the interdecadal variations...
of the EAWM and their association with the quasi-stationary planetary wave activity. The EAWM is shown to be strong during 1976–87 and it experienced a significant weakening around 1988. This change has an important influence on the East Asian winter climate, which is characterized by warming over large areas of the East Asia after 1988. The frequency of occurrence of cold waves in China also decreased significantly.

To investigate the internal dynamical process for the interdecadal variation of EAWM in the late 1980s, the associated quasi-stationary planetary wave activity, including the amplitudes and EP fluxes for zonal wave-numbers 1 to 3, are analyzed by using the ERA-40 and NCEP–NCAR reanalysis datasets. By defining a PWA index, it is shown that the activity of the quasi-stationary planetary wave clearly had an interdecadal variation around 1988. Compared with the situation during 1976–87, the propagation of quasi-stationary planetary waves from the troposphere into the stratosphere along the polar waveguide became weak during 1988–2001, whereas the propagation to the upper troposphere over the subtropics along the low-latitude waveguide was obviously enhanced. This leads to the convergence of EP fluxes around 35°N and decelerates the zonal-mean zonal winds in this region. The East Asian jet stream is then weakened, which indicates weakening of the EAWM. An interesting feature is that the decadal changes of zonal-mean zonal wind are mainly in the troposphere, implying that such change may mainly come from the lower part of the atmosphere.

In addition to the changes in the wave-convergence-induced jet stream, the changes in the amplitude of planetary waves also contribute to weakening of the EAWM. The amplitude of planetary waves is significantly weakened around 45°N since 1988, with the anomaly center extending to the surface. This change in the wave amplitude weakens both the Siberian high and Aleutian low. Therefore, the pressure gradient between the two systems is reduced and the EAWM is weakened. Further analyses indicate that the zonal wavenumber 2 pattern of planetary waves plays the most important role in this process. The wavenumber 2 pattern coincides well with climatological pressure systems, including the Siberian high and Aleutian low. It accounts for more than 75% and 50% of the interdecadal changes in the Siberian high and the Aleutian low after 1988, respectively, and therefore contributes dominantly to the weakening of the EAWM. More evidence reveals that this decadal weakening of wavenumber 2 may be induced by the change of land–sea thermal contrast; during this process the decrease of snow depth over the Eurasian continent and North America plays an important role. It is also worth noting that the reason for land–sea thermal contrast

FIG. 13. Patterns of the zonal wavenumber 2 distribution of winter mean SLP for the winters of (a) 1976–87 and (b) 1988–2001 and (c) their difference (1988–2001 minus 1976–87). Contour intervals are 2 hPa in (a),(b) and 0.5 hPa in (c). Light shading in (c) indicates the 95% confidence level.
change is complex, and many factors such as SST, sea ice, and soil moisture may also be involved. Their roles need to be investigated in the future.

The analyses of this study are based on the winter means. In fact, the decadal variations of EAWM can also be revealed by changes in the frequency of strong and weak EAWM months. Such analyses may also be helpful to understand the underlying dynamics and therefore worth doing in the future. Another point that should be noted is, in this study, only the internal dynamical variations of quasi-stationary planetary waves and their effects upon the EAWM on the interdecadal time scale are investigated. The reason why the planetary wave varies on the interdecadal time scale is not fully discussed. On the decadal time scale, the external forcing or a coupling forcing, such as ocean or land processes, may be the controlling factor (e.g., Watanabe and Nitta 1999), and feedbacks may exist between these factors and the internal atmospheric processes. To understand this feedback process, analyses based on data with higher temporal resolution (e.g., monthly mean) or on a coupled atmospheric and oceanic general circulation model are needed in the future.

Acknowledgments. We thank the two anonymous reviewers for their valuable comments and suggestions in improving the quality of the paper, and appreciate the European Centre for Medium-Range Weather Forecasts and the National Centers for Environmental Prediction–National Center for Atmospheric Research for providing their reanalysis datasets. This work is supported jointly by the National Key Technology R&D Program of China (Grant 2008BAK50B02), the National Basic Research Program of China (Grant 2009CB421405), and the National Natural Science Foundation of China (Grant 40730952).

REFERENCES


